PUBLIC HEALTH IMPLICATIONS OF AIRBORNE INFECTION: PHYSICAL ASPECTS

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The term airborne infection implies a physical transport of the causative agent by the atmosphere from the source to the receptor. Although primary concern from a public health standpoint has been confined to the transport within a single room or building, transport in the free air outside a confining structure cannot be overlooked. Open-air transport has in fact been referred to at this Conference by Dr. Smith in connection with his discussion of the occurrence of *Coccidioides immitis* infection during the construction of airfields in the San Joaquin Valley. In this case airborne transport was probably substantially less than 1 mile. In a recent unpublished report by C. E. Wellock and M. F. Parker, an airborne infection of *Coxiella burnetii* is described involving organism transport from a rendering plant to a distance of 7½ miles in an urban area. The transport of organisms to much greater distances, i.e., several hundreds of miles, has been discussed by Jacobs (4) and is a well-recognized phenomenon in connection with epiphytotics such as those occurring with stem rust of wheat in the central United States. Whether such large-scale transport is a public health problem depends primarily on the source strength and the viable decay rate.

The purpose of this paper is to provide a brief nonmathematical discussion of the physical factors controlling the behavior of an aerosol as it is carried downwind by the atmosphere from an established source. These factors include physical losses, i.e., fallout, impaction on surfaces, and washout by precipitation, together with dilution resulting from atmospheric diffusion processes. All of these factors taken together will determine the change in airborne concentration or dosage while the aerosol cloud is being transported. Loss in viability during the transport period is specifically excluded from consideration since this is a biological rather than physical factor and is a specific characteristic of the airborne material, whereas the aforementioned factors are not.

It is assumed that the aerosols of interest have particle densities near 1 g/ml and that the particle diameters are predominantly 1 μ with a maximum of 5 μ. For illustrative purposes downwind travel is considered in three categories, viz., short, intermediate, and long distances corresponding to a few hundred yards, approximately 10 miles, and greater than 100 miles, respectively. Physical losses and diffusive processes appropriate to aerosols in the 1- to 5-μ range are examined separately in the following sections.

Physical Losses

For particles in the size range of interest, it can be shown that physical losses play a very minor part in affecting the concentration or dosage at ground level under most circumstances. There are reasonably good theoretical grounds for estimating aerosol losses resulting from fallout and washout; impaction losses can be estimated only very roughly.

Fallout loss. The quiescent terminal settling velocity, v_s, is proportional to the quantity ($d^2\rho$) where d is the particle diameter and ρ the particle density. For 1- and 5-μ particles of unit density, terminal velocities are 0.03 and 0.76 mm/sec or 0.11 and 2.7 m/hr, respectively. Under all but the most unusual atmospheric conditions an aerosol cloud released near the ground is subject to vertical mixing velocities much greater than these settling velocities. Consequently, fallout takes place under turbulent rather than quiescent conditions and, at the same time, the aerosol cloud is mixed vertically thereby increasing the height particles must fall to reach the ground. Thus, both turbulent settling and vertical cloud mixing reduce the fallout loss relative to that expected based simply on the quiescent terminal velocity and the initial cloud height.

To obtain a quantitative estimate of the importance of fallout in reducing the airborne concentration or dosage near the surface, it is convenient to make two simplifying assumptions regarding the behavior of the aerosol in the atmosphere. First, as a result of vertical mixing, the surface dosage with or without fallout de-

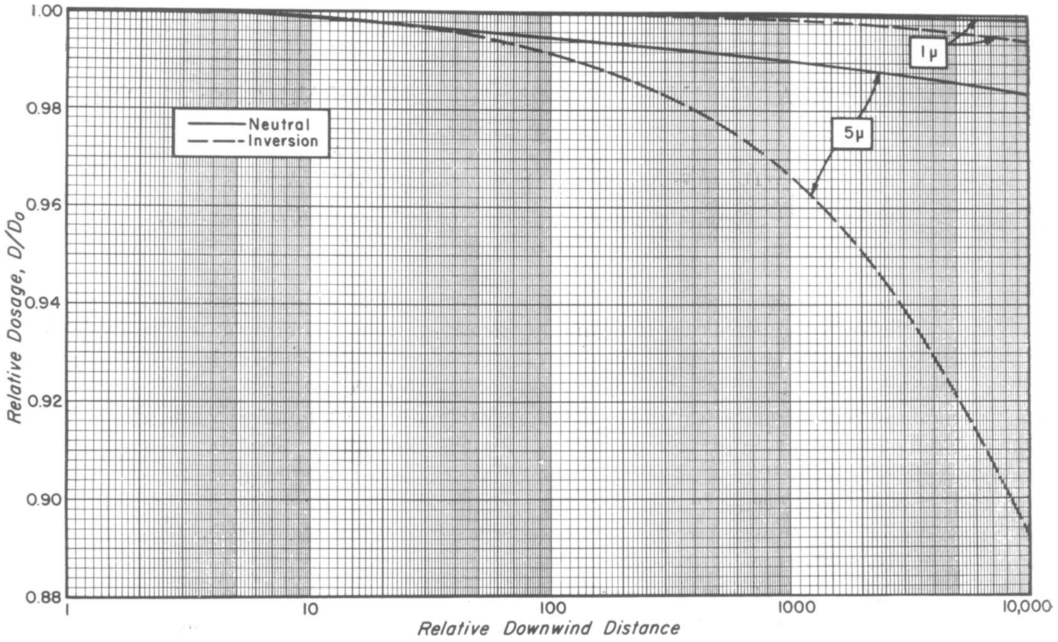

FIG. 1. *Effect of fallout on downwind dosage. Ratio of dosage with and without fallout, D/D_0, under neutral and inversion conditions for 1- and 5-μ particles of unit-density as a function of downwind distance. Distance is given relative to the cloud height when deposition begins. (Mean wind speed, 10 miles/hr.)*

creases inversely as a power of the downwind distance from the source. Second, the surface deposition in terms of particles per unit area is equal to the product of the terminal settling velocity and the surface dosage.

Based on these assumptions, curves are given in Fig. 1 showing the calculated ratio of surface dosage (or concentration) without fallout to that with fallout as a function of downwind distance for two particle sizes and for conditions appropriate to minimal vertical mixing (inversion conditions) and moderate vertical mixing (neutral conditions). To show the results in most general form, the downwind distance is given relative to the vertical height of the aerosol cloud at the time the bottom of the cloud first reaches the ground. As expected, fallout loss is greatest when vertical mixing is at a minimum. For example, an aerosol cloud of 5-μ particles extending from the ground to 1 m height will have lost approximately 8% of its particles after traveling 5,000 m at 5 m/sec (10 miles/hr) and the relative dosage at that distance will be 0.92. The corresponding dosage ratio for an aerosol cloud of 1-μ particles is 0.996. A reduction in wind speed results in a proportionate reduction in distance to which a given dosage ratio is applicable.

An examination of Fig. 1 shows that the expected fallout loss for 1-μ particles is completely negligible and that the loss for 5-μ particles is not significant except, perhaps, for very large travel distances under conditions of minimal vertical mixing.

Washout loss. Airborne particles are subject to capture and removal by falling raindrops. The process involves impaction of the particles by the drops and takes place with an efficiency that can be estimated from theoretical and experimental considerations Obviously, the total loss of particles by washout will depend on the impaction efficiency and the nature of the precipitation.

In the most recent investigation of the washout problem, Greenfield (3) shows that impaction loss depends primarily on the total rainfall to which the aerosol cloud is exposed rather than the rainfall intensity or rate. Based on Greenfield's estimates, including appropriate values of the impaction efficiency, curves are given in Fig. 2 showing the fraction of airborne particles remaining airborne after exposure to various amounts of rainfall during cloud travel. Evidently the loss is extremely dependent on particle size; 0.6 in. of rain may remove as much as 90% of the 5-μ particles but only 7% of the 3-μ particles. How-

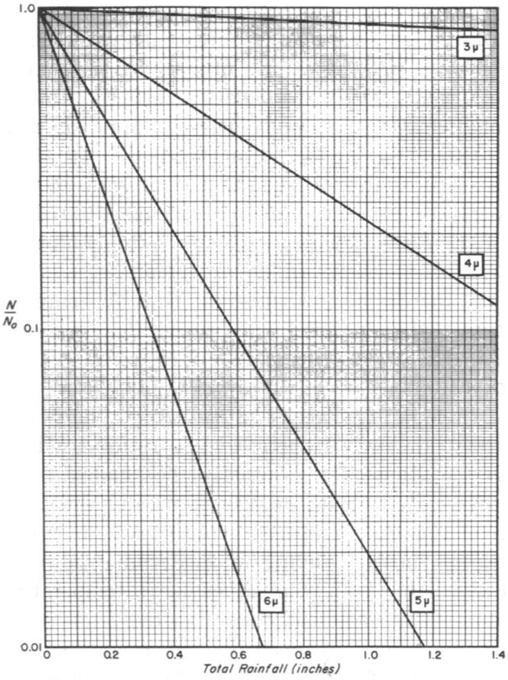

FIG. 2. *Physical loss by precipitation. Fraction of airborne particles remaining, N/N_0, after exposure to various amounts of rain.*

ever, assuming that 5 μ represents the upper limit and that most of the particles of interest are in sizes less than 3 μ, rain washout cannot be considered an important factor for small- or intermediate-scale travel or even large-scale travel unless the rainfall is very great.

Impaction loss. Loss of particles by impaction on obstructions in the path of the moving cloud depends on the impaction efficiency and the number of times the entire particulate cloud encounters the obstructions. It is well known that impaction efficiency of airborne particles increases with increasing values of a dimensionless parameter, K, defined by

$$K = \frac{1}{18} \frac{\rho d^2 V}{\eta D}$$

where ρ is the particle density; d, the particle diameter; V, the velocity of the particle relative to the collector; D, the minimal crosswind dimension of the collector; and η, the viscosity of air.

Langmuir (5) has shown that if the numerical value of K is less than 0.07, the theoretical impaction efficiency is zero. Under conditions appropriate to this K value, particles will move past the obstruction without making contact and as a consequence, there will be no loss on the obstructing surface. For example, if the obstruction is 1 cm across and 5-μ particles approach at 10 miles/hr (500 cm/sec), $K = 0.04$, which is well below the limit required for finite impaction loss. This suggests that only the smallest surfaces such as foliage and wires will be at all effective in removing 5-μ particles from the atmosphere. Obviously, for given conditions, K values for 1-μ particles are smaller than those for 5 μ by a factor of 25 and impaction loss, if finite, is also smaller by a corresponding amount.

To obtain significant losses even of the larger particles, the obstructions must be dense enough to provide repeated chances for interception, as would be the case in a thick forest or tall grass. However, under these conditions the wind speed will be reduced and the K value and the efficiency will decrease accordingly. If the obstructions are more widely separated so that the wind speed can be maintained, vertical mixing will distribute the particles aloft and only a portion of the cloud will remain at levels where contact can be made. Thus, the conditions necessary to achieve significant loss by impaction, namely strong winds, a high density of obstructing surfaces and restricted cloud height, are, to a degree, mutually exclusive.

Unfortunately, except for travel over urban areas, test data are not available to give quantitative estimates of impaction loss, but these qualitative considerations indicate that the losses are not a major factor in reducing aerosol dosages. Surface dosages measured within urban areas using a fluorescent particulate tracer show that both fallout and impaction losses are negligible, as will be pointed out in the following section.

Dilution by Atmospheric Diffusion

To the extent that physical losses may be neglected, dilution of a particulate cloud as it moves downwind is identical to that of a gaseous cloud. Accordingly, data on diffusion of gases can be applied to aerosols in the 1- to 5-μ size range.

The diffusive process is dependent on meteorological conditions and the nature of the terrain over which the airborne cloud passes. Mixing within the atmosphere results from the interaction of parcels of air or eddies moving slower or faster than the mean wind and in directions different from the mean wind direction. The eddies may be present in a wide range of sizes. In general those much smaller than the aerosol cloud cause internal mixing and hence tend to produce a uniform distribution of particles within the cloud. Those much larger than the cloud displace the cloud bodily with respect to its mean direction and are recognized locally as causing significant changes in wind direction. Eddies comparable in size to the aerosol cloud are most effective in exchanging outside air with the aerosol, causing an increase in cloud size and consequent dilution of the airborne material. The mixing or diffusion process takes place vertically and horizontally. However, to simplify the discussion as presented here for both small- and intermediate-scale travel, horizontal mixing normal to the mean wind direction can be eliminated by considering cloud travel from extended crosswind line sources only. With this limitation, lateral mixing will not result in aerosol dilution. Further, if one is interested only in total dosage[1] at a downwind position, mixing and dilution in the direction of cloud travel are eliminated. Consequently, the change in dosage with distance from a long line source depends only on the extent of dilution resulting from vertical mixing.

In general, the thermal stability of the atmosphere as indicated by the vertical temperature gradient is the simplest and most easily understood index of the relative degree of vertical mixing. For convenience, three stability categories may be considered, viz., inversion, neutral, and lapse in order of decreasing stability or increasing vertical mixing. The difference between these three categories may be illustrated by the behavior of an idealized parcel of air transported vertically without mixing and without gain or loss of heat, i.e., adiabatically. Since the atmosphere is a compressible gas, adiabatic compression of such a parcel will cause a temperature increase and adiabatic expansion, a temperature decrease. By virtue of the decrease in atmospheric pressure with height, the temperature of the parcel of air must decrease if carried upward by an amount appropriate to the adiabatic cooling. The adiabatic temperature change is approximately constant for the lowest few thousand feet of atmosphere and is 0.5 F/100 ft. This vertical temperature gradient applies only to unsaturated air. If there is condensation of water, the gradient is less (0.3 F/100 ft). In this discussion it is assumed that there is no condensation. Thus the temperature of the air parcel will decrease 0.5 F for each 100 ft it is carried up and will increase 0.5 F for each 100 ft it is carried down.

Under conditions of neutral atmospheric stability, the actual vertical temperature gradient in the atmosphere is exactly the adiabatic gradient. This means that a vertically displaced parcel of air will always have the same temperature as the surrounding air. Under inversion conditions, however, the atmospheric temperature gradient is such that there is a temperature increase rather than decrease with height. In this case, a parcel lifted adiabatically will always be cooler and therefore denser than the surrounding air and will tend to subside to its original position. Similarly, if carried downward, the parcel will be warmer and less dense than the air around it and will therefore rise to its original level. Finally, under lapse conditions, the actual temperature gradient in the atmosphere is such that the temperature decreases more rapidly with height than does the adiabatic rate. As a result, a parcel displaced upward will be warmer and less dense than the surrounding air; hence its upward motion will tend to continue. If displaced downward, the parcel will be cooler and again the motion will continue.

Vertical mixing is to be expected as a result of turbulence generated by air moving over the ground surface. Under inversion conditions, the turbulence and vertical mixing will tend to be damped out. Under lapse conditions, vertical mixing will be enhanced and in fact may be reinforced by thermal convection. Neutral conditions represent an intermediate degree of vertical mixing.

In the surface layer, i.e., the lowest 100 ft or so, inversion conditions develop at night particularly if the sky is clear so that the ground surface can cool by radiation. Lapse conditions develop during the day if there is surface heating. However, both of these extremes in atmospheric

[1] Dosage is defined as the total number of particles collected at unit flow rate or breathing rate and is numerically equal to the total number of particles collected divided by the flow rate. It is also numerically equal to the product of average concentration and exposure time. Dosage is to be distinguished from dose which is defined as the total number of particles collected at a specified flow rate.

stability may be modified by surface irregularities and moderate wind speed. Neutral conditions will obtain if the surface is neither heated nor cooled as may be the case if the sky is overcast.

If a deeper layer is considered, i.e., 1,000 ft or more, the temperature gradient may be variable. Inversions aloft with neutral or unstable air below are not uncommon as a result of either subsidence or the horizontal advection of warm air above or cool air below. Obviously, the height to which the temperature structure of the air must be considered is greater the larger the scale of cloud travel.

Small-scale travel. Both experimentally and theoretically, greatest attention has been given to the problem of small-scale travel. Theoretical aspects are beyond the scope of the present discussion. A comprehensive treatment of atmospheric turbulence has been given by Sutton (12) and more recently by Priestley (9). Representative examples of experimental results are given in Fig. 3 to illustrate the effect of atmospheric

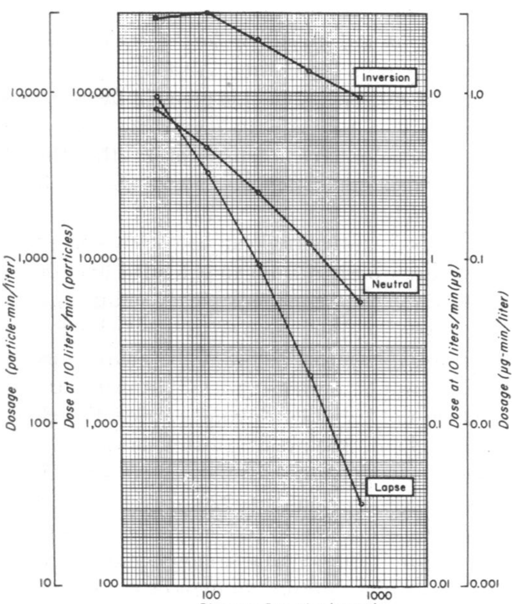

FIG. 3. *Change in dosage or dose with downwind distance from a long line source under lapse, neutral, and inversion conditions. Data are based on O'Neill SO_2 field tests no. 16 (lapse), 22 (neutral), and 32 (inversion) adjusted to a line source of either 10^{10} particles/meter or 1 g/meter. Use left-hand scale for source in particles and right-hand scale for source in grams.*

stability on the dilution of airborne material transported to a distance of approximately 800 m. The three sets of data, taken from a comprehensive series of field tests run at O'Neill, Nebraska, in 1956, as reported by Barad (1), show the marked difference in rate of change of dosage with distance under inversion, neutral, and lapse conditions over smooth, flat terrain. In all of the O'Neill trials, sulfur dioxide was released from a point source ½ m above ground and sampled at 1 m or higher on concentric arcs downwind from the source. For presentation in Fig. 3, computed crosswind integrated dosages from the point sources are converted to equivalent axial dosages from long line sources. Dosages have been adjusted to a common source strength for each trial. For illustrative purposes both dosage (quantity sampled at unit flow rate, i.e., 1 liter/min) and dose (quantity sampled at 10 liters/min) are given for each sampling distance. The dosage and dose are shown in terms of weight for a source of 1.0 g/m and also in terms of numbers of particles for an aerosol cloud having an arbitrary source strength of 10^{10} particles/m. A straight line in Fig. 3 indicates that the dosage (or dose) varies with distance, X, as $1/X^{\beta}$, where β is an experimentally determined exponent. Values of β and of the ratio of dosages at 100 and 800 m for the three trials are summarized below. The dosage ratio as given indicates the relative degree of dilution under the three conditions.

	Inversion	*Neutral*	*Lapse*
Dosage at 100 m/dosage at 800 m	3.3	8.5	102
β	0.57	0.96	2.0

The above values apply to specific trials, but they are representative of what one might expect over flat, open terrain under the three conditions of atmospheric stability. The marked effect on surface dosage resulting from different degrees of vertical mixing associated with the three categories of atmospheric stability is clearly evident from the dosage ratios given above and the plotted data in Fig. 3.

Intermediate-scale travel. In traveling a distance of the order of 10 miles, an aerosol (or gas) cloud may become distributed vertically through a distance of several hundred or even a few thousand feet. Within this height the temperature structure may be quite irregular. For example, daytime lapse conditions near the surface may be accompanied by relatively stable inversion conditions aloft. This combination is frequently observed along the west coast of the United States and is often found in other parts of the country as well. Evidence for this condition is a sharp top to a visible haze layer usually well distributed below its upper boundary. The night-time temperature gradient over a city is of special interest. Duckworth and Sandberg (2) have shown clearly that nocturnal surface inversions do not develop within urban areas even though stable inversion conditions prevail in the surrounding open areas. Heat retained within the city gives rise to more nearly neutral conditions within a layer equivalent to two or three times the average building height. Above this height the temperature gradient may be quite irregular.

In general, vertical mixing aloft is less than that at the surface; hence, on an intermediate scale, surface dosages do not decrease with downwind distance as rapidly as is suggested by the data from small-scale trials. This point is well illustrated by trials conducted over the city of Los Angeles using a fluorescent particle tracer technique developed by the Aerosol Laboratory, Stanford University. The technique is based on the fact that 1- to 5-μ fluorescent particles can be released in known numbers and sampled quantitatively. Since the tracer particles are easily distinguished from other particles in the atmosphere and can be counted individually, the method is capable of giving high sensitivity and is readily adapted to intermediate-scale travel.

Results from one of the daytime trials conducted by Neiburger (8) are shown in Fig. 4. An aerosol cloud, containing 1.0×10^{14} tracer particles (2,400 g) with a mass mean diameter of 2.3 μ, was released from a point source in the early morning. Light winds carried the tracer first to the southeast, then north. Shortly after 11:00 AM the wind speed increased, and between 12:00 and 1:00 PM the main portion of the aerosol cloud crossed the sampling arc 10 miles distant from the source. Wind speed at the time of crossing was about 8 miles/hr and the wind direction was 45° to the sampling line. The approximate trajectory is shown in Fig. 4 together with the total dosages observed at the various sampling stations.

It is important to note that during the cloud travel period the lower atmosphere was unstable due to lapse conditions from the surface to 300 m. At this height the temperature gradient changed to inversion conditions indicating stability aloft. As a result there was substantial vertical mixing below 300 m but little or no mixing above this height. The mixing process would, of course, tend to produce a uniform distribution of the tracer particles from the surface to 300 m. Thus one would expect dosage to be constant with height in the sense that the sum of all dosages observed at ground level is the same as that expected at other levels up to 300 m. Accordingly, since the cloud was bounded laterally, the surface dosages, mean wind speed, and cloud height can be combined to estimate the total number of particles passing through a 300-m high vertical plane at the sampling arc. The total number of particles, Q, passing through a vertical plane of height H at mean wind speed U normal to the plane is $Q = HU \int_0^\infty D dy$. The $\int_0^\infty D dy$ is the crosswind integrated dosage. The computed number in this trial is 1.1×10^{14} particles, which is in excellent agreement with the 1.0×10^{14} particles released. The close agreement between the computed and known number of airborne particles suggests that both fallout and impaction losses were negligible during the 6-hr travel time. Although this fact is in accord with the discussion given in the previous section, it is particularly significant since the tracer particles have a density of 4 rather than 1 g/ml, which would tend to increase both fallout and impaction.

Results from this trial can be readily converted to dosages expected from an extended line source under these same meteorological conditions. If a long crosswind line source of 10^{10} particles/m had been used (approximately 0.24 g/m), surface dosages would decrease with downwind travel distance until vertical mixing carried the cloud to a height of 300 m. At greater downwind distances, surface dosages would remain essentially constant at a value of 220-particle min/liter (assuming wind speed and inversion height are unchanged). Thus, if the sampling rate were 10 liters/min, the dose or total recovery would be 2,200 particles. By way of contrast, had vertical mixing continued throughout the 10-mile travel distance and at the rate found for the small-scale lapse trial, the computed surface dosage would have been two orders of magnitude smaller than that given above.

Large-scale travel. It is well known that naturally occurring aerosols of dust and spores travel

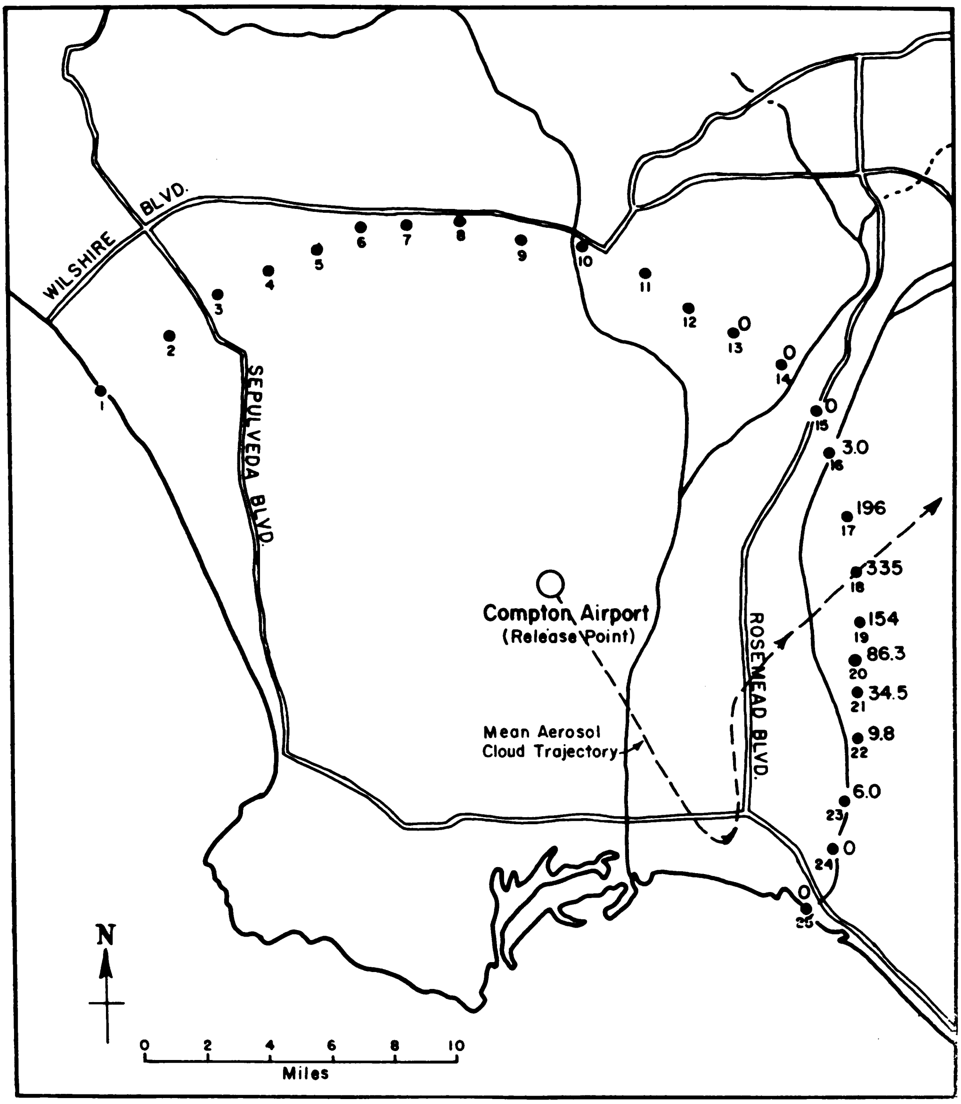

FIG. 4. *Intermediate-scale travel in Los Angeles. Total dosages in particle minutes per liter obtained from a point source of 1.0 × 10*^{14} *fluorescent particles (2,400 g). Cloud passed sampling line at approximately 12:00 hours with mean wind speed of 8 miles/hr. (Trial 5, Air Pollution Foundation Report No. 7; Neiburger (8).)*

hundreds or even thousands of miles. For example, Mill and Lempfert (6) examined dust which appeared during a storm over southern England and northern Europe and determined from its physical characteristics that it must have originated in the Sahara Desert. His wind trajectory analysis verified the source location and suggested that the cloud followed a circuitous path over the Atlantic for a distance of some 2,000 miles. Dust reported by ships in the proposed trajectory confirmed his meteorological analysis. It is interesting to note that Mill reported particle sizes to be 1 to 10 μ in diameter with a maximal number at 6 μ.

Dust storms in the United States also demonstrate a close correlation between meteorological conditions and the transport of material. Miller (7) showed that a storm in South Dakota generated a dust cloud which moved southeast to the Atlantic coast. An examination of the meteorological records showed that the cloud direction and time of travel coincided with a frontal system moving across the eastern United States. Another dust cloud originating in northern Texas caused discolored rain and snow in New England 24 hr later. Robinson (10) examined the dust combined with the precipitation and found a maximal number of particles in the 6- to 20-μ range with a chemical composition characteristic of the source area. Again the direction of cloud travel agreed well with the meteorological trajectory. For several hundred miles the dust was reported between 6,000 and 10,000 ft but not on the ground. During this time the southerly air was being carried over a wedge of cold air moving slowly from the northwest so that the apparently anomalous behavior of the cloud was in accord with the known meteorological conditions.

Airborne transport of the urediniospore responsible for stem rust of wheat has long been recognized as a critical problem in the central United States. Every few years there is a coincidence of favorable conditions of moisture, temperature, stage of plant growth, and spore transport with disastrous results. For example, during the first week of June, 1925, spores were carried north from Oklahoma and southern Kansas by a general circulation pattern associated with the Bermuda high pressure system. Stakman (11) reported that severe infection broke out 10 days later on a 400-mile-wide front to a distance 600 miles downwind from the source. The area infected agreed well with the wind trajectories during the transport period. Thus even though these spores are relatively large, approximately 15 by 20 μ, they were carried 600 miles in sufficient amounts to produce serious crop loss.

There is ample evidence, therefore, that large particles are transported hundreds of miles; hence, there is little doubt that 1- to 5-μ particles can also be transported over great distances. From the evidence at hand it is apparent that synoptic meteorological data can be used to estimate the horizontal direction and speed as well as general vertical motion of the aerosol during large scale transport of airborne material. Obviously, however, it is impossible to ascertain the source strength for these naturally occurring aerosols; and, unfortunately, reliable sampling data giving dosage or concentrations are virtually nonexistent. Hence, little can be said regarding quantitative aspects of the diffusive properties of the atmosphere for large-scale transport of aerosol.

Summary

The physical aspects of the transport of unit-density, 1- to 5-μ aerosol particles in the free atmosphere, are considered for small-, medium-, and large-scale travel. Losses resulting from fallout, washout by rain, or impaction on surfaces do not significantly affect dosage or concentration. There is a substantial amount of theoretical and experimental information on the diffusion of airborne material to distances of a few hundred yards (small scale) under inversion, neutral, and lapse conditions. Only limited quantitative data are available for travel distance the order of 10 miles. On this intermediate scale, dilution by diffusion processes may be less than expected from an extrapolation of the small-scale results. Large-scale transport for hundreds of miles is entirely possible. Direction, speed, and certain aspects of the vertical motion can be deduced from synoptic meteorological information, but there is little or no information regarding quantitative aspects of the diffusion processes on this scale.

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